Thermal localization as a potential mechanism to rift cratons

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A B S T R A C T

Cratons are cold regions of continents that have remained stable since at least the Precambrian. The longevity of cratons is often attributed to chemical buoyancy and/or high viscosity of cratonic root material. Yet examples of destructed cratons (such as the North China Craton) suggest that there are conditions under which chemical buoyancy and/or high viscosity are insufficient to keep cratons stable. The formation of continental rifts, as weak zones that reduce the total stresses of cratons, may be a mechanism that increases the longevity of cratonic lithosphere. Since continental rifts result from localized deformation, understanding the mechanism of shear localization is thus important for understanding the stability or breakup of cratons. Here, we perform 2-D numerical models for a cratonic lithosphere under extension to understand the initiation of shear localization, for visco-elastic-plastic rheologies. Results reveal that three modes of deformation exist: no localization, symmetric localization and asymmetric localization. To further understand the underlying physics, we develop a 1-D semi-analytical method that predicts the onset of localization as well as whether rifting will be symmetric or asymmetric. Applications of the semi-analytical method to geological settings show that fast deformation, cold thermal state or strong mantle rheology may result in localized deformation and stabilize the remaining adjacent cratons. Our results successfully interpret the major features of the North China Craton.

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1. Introduction

Archean cratons are regions of continents that have remained undeformed since at least the Precambrian. They are characterized by relatively low surface heat flow, diamond deposits in kimberlites, and thick keels with high seismic velocities. Most Archean cratons have undergone minimal internal deformation, despite being directly adjacent to orogens or rifts (e.g. the east African Rift System, the Baikal Rift) at their boundaries (e.g. Stern, 1994; O’Neill et al., 2008; Sengor, 1999).

Yet, there are examples showing that cratonic lithosphere can be destructed as well, e.g. the Wyoming Craton and the North China Craton (NCC) (Eggler et al., 1988; Carlson et al., 2004; Fan and Menzies, 1992; Menzies et al., 1993; Griffin et al., 1998).

Although the longevity of cratonic lithosphere is often attributed to chemical buoyancy and/or high viscosity of cratonic root material (Shapiro et al., 1999; Sengor, 1999), they are relatively ineffective if cratons are involved with subduction zones (Lenardic et al., 2000, 2003). The existence of weak mobile zones surrounding cratons, in addition, was proposed to be a possible mechanism to increase the longevity of cratonic lithosphere (Lenardic et al., 2000). Such weak zones could be continental rifts and orogens that have relatively low yield stresses. Once a weak zone is initiated, the integrated stress of the lithosphere would reduce significantly which keeps the remaining part of the craton relatively undeformed. This might explain why most cratons are stable, even though they are adjacent to rifts or orogens (O’Neill et al., 2008). The formation of continental rifts is due to lithospheric runaway in which deformation is strongly localized in narrow zones (e.g. England, 1983; Buck, 1991; Buck et al., 1999). While some authors (e.g. England, 1983; Buck, 1991) extended the concept of rifts to be narrow rifts or wide rifts, it is the narrow rifts that behave as weak zones and stabilize cratons. Wide rifts, in contrast, can be observed in destructed cratons, e.g. the eastern North China Craton (Zhao and Xue, 2010), and the West Siberia Basin (Armitage and Allen, 2010). They are thought to result from stretching of relatively thick continental lithosphere without shear localization (Armitage and Allen, 2010; Buck, 1991). Better understanding the mechanisms of localization is thus important to understand dynamics of continental evolution.

A number of studies have been focusing on the mechanisms of localization (e.g. Bercovici and Karato, 2003; Montesi and Zuber, 2002; and references therein). Dynamic mechanisms of self-softening are required to generate localization (Bercovici and Karato, 2003; Montesi and Zuber, 2002). Proposed mechanisms include...
The compressible Boussinesq equations for a slowly deforming lithosphere are given by
\[
\begin{align*}
\frac{\partial \mathbf{v}}{\partial t} & = 0 \\
\frac{\partial \sigma_{ij}}{\partial x} + \rho_0(1 - \alpha(T - T_0))g_i & = 0 \\
\rho C_p \left( \frac{\partial T}{\partial t} + v \frac{\partial T}{\partial x} \right) & = \frac{\partial}{\partial x} \left( k \frac{\partial T}{\partial x} \right) + H_r + H_t
\end{align*}
\]
where \( v_i \) is the velocity, \( x_i \) is the spatial coordinates, \( t \) is the time, \( T \) is the temperature, \( \rho \) is the density, \( \alpha \) is the thermal expansivity, \( g \) is the gravitational acceleration, \( \sigma_{ij} = -P \delta_{ij} + \tau_{ij} \) is the total stress, \( P = -\sigma_{ij}/3 \) is the pressure, \( \delta_{ij} \) is the Kronecker delta, \( \tau_{ij} \) is the deviatoric stress, \( C_p \) is the heat capacity, \( k \) is the thermal conductivity, \( H_r \) is the radioactive heat production, and \( H_t = \tau_{ij} \left( \delta_{ij} - \frac{1}{3} \delta_{ij} \right) \) is the shear heating.

The strain rate is defined as
\[
\dot{\varepsilon}_{ij} = \frac{1}{2} \left( \frac{\partial v_i}{\partial x_j} + \frac{\partial v_j}{\partial x_i} \right)
\]

The rheology of rocks is Maxwell visco-elasto-plastic:
\[
\begin{align*}
\tau_{ij} &= \frac{\dot{\varepsilon}_{ij}}{2 \mu} + \frac{1}{2G} \frac{D\tau_{ij}}{Dt} + \lambda \frac{\partial Q}{\partial \sigma_{ij}} \\
D\tau_{ij} &= \frac{\partial \tau_{ij}}{\partial t} + v_k \frac{\partial \tau_{ij}}{\partial x_k} - W_{ik} \tau_{ij} + \tau_{ik} W_{kj}
\end{align*}
\]
where \( \mu \) is the effective viscosity, \( G \) is the elastic shear module, \( \lambda \) is the plastic multiplier, \( Q \) is the plastic flow potential and \( W_{ij} = \frac{1}{2} \left( \frac{\partial s_{ij}}{\partial x} - \frac{\partial s_{ij}}{\partial x} \right) \) is the vorticity.

The creep rheology is strain rate and temperature dependent:
\[
\mu = A \lambda^{-1} \exp \left( \frac{E}{nRT} \right)
\]
where \( A \) is a pre-exponential parameter, \( n \) is the power-law exponent, \( \lambda = (0.5 \mu / E)^{1/2} \) is the second invariant of the strain-rate tensor, \( E \) is the activation energy and \( R \) is the universal gas constant.

The brittle strength of crustal rocks is approximated by Mohr–Coulomb plasticity:
\[
\begin{align*}
\tau &= \tau' - \sigma' \sin(\phi) - C \cos(\phi) \\
Q &= \tau'^2
\end{align*}
\]
where \( \tau' = ((0.5(\sigma_{xx} - \sigma_{xx}))^{2} + \sigma_{zz}^{2})^{1/2} \), \( \sigma' = -0.5(\sigma_{xx} + \sigma_{xx}) \). \( C \) is cohesion and \( \phi \) is the internal angle of friction. For rocks in the mantle, the yield strength is described by “Peierls” or “low-temperature” plasticity. Here, the formulation of Goetze and Evans (1979) is applied which is valid for differential stresses that are larger than 200 MPa:
\[
\tau_{ij} = \frac{\dot{\varepsilon}_{ij} \sigma_0}{\sqrt{3} \lambda} \left( 1 - \frac{RT}{H_0} \ln \left( \frac{\sqrt{3} E_0}{2 \lambda} \right) \right)
\]
where \( \sigma_0 = 8.5 \times 10^9 \) Pa, \( E_0 = 5.7 \times 10^{11} \) s⁻¹ and \( H_0 = 525 \) kJ/mol.

2.2. The 2-D model

Our standard 2-D setup is 800 km wide by 670 km deep (Fig. 1). The model is laterally homogeneous and consists of an upper crust (25 km), a lower crust (15 km), a mantle lithosphere (150 km) and asthenosphere mantle (480 km) (see Table 1 for model parameters). The whole model with a free surface is subjected to pure shear extension with prescribed extensional strain rate \( \dot{\varepsilon}_{bg} \). The thermal boundary conditions are zero-flux at side boundaries and isothermal at the top (0°C) and the bottom (1600 °C) boundaries. The steady-state initial temperature is obtained by performing half-space cooling from 1600 to 0 °C at the surface over a time thermal age. The deep part of the mode, where temperature is greater than 1300 °C, is replaced by an adiabatic temperature gradient of 0.5 °C/km. Deviatoric stresses are initially zero. The extensional strain rate is varied from 10⁻¹⁶ to 10⁻¹³ s⁻¹. The initial thermal age, which indicates temperature of the lithosphere by conduction only, is varied from 50 to 2000 Ma. It should be noted that the thermal age is typically different from geological ages. Due to convection in the real Earth, the lithosphere does not cool down significantly after about 600 Ma of geological age. Yet, as we would like to do systematic simulations and obtain the onset of localization versus lithosphere strength and extension strain rate, we still perform simulations with these parameters although they are somewhat unrealistic. Random noise is present in the models because the locations of the tracers are initially randomly perturbed.
Tests with prescribed small weak seeds reveal that such weak seeds would focus strain localization and result in an earlier occurrence of localization. No additional thermal or rheological heterogeneities are included in the model.

The 2-D simulations are done with the finite element code MILAMIN_VEP (Kaus et al., 2010; Crameri and Kaus, 2010). MILAMIN_VEP solves a visco-elasto-plastic problem in a purely Lagrangian manner, and uses remeshing to deal with large deformations. The crust has a Mohr–Coulomb plasticity and the mantle has a low-temperature (Peierls) plasticity (Goetze and Evans, 1979). Shear heating is included. The code has been benchmarked and used in a number of previous studies (e.g. Kaus et al., 2009; Schmeling et al., 2008; Schmalholz et al., 2009; Crameri and Kaus, 2010). The resolution in this work is set to be $201^2$ for the mesh grids. Tests with higher resolution indicate that increasing the resolution yields narrower shear zones but does not alter the overall results significantly. Around 0.2 million randomly distributed markers are used. We find that perturbation introduced by the randomness of markers is sufficient to initiate not only symmetric but also asymmetric localizations.

![Fig. 1. (Left) Model setup for 2-D simulations. The model consists of an upper crust (25 km), a lower crust (15 km), a lithospheric (150 km) and an asthenospheric (480 km) mantle. The whole model has a constant strain rate condition at the bottom and side boundaries, and a free surface boundary condition on the top. Thermal boundary conditions are isotherm at the top ($T = 0$) and bottom ($T = 1600$) boundaries and zero heat flux at side boundaries. (Right) Illustration of initial temperature and strength profile for the standard model with a thermal age of 500 Ma under an extension strain rate of $10^{-14}$ s$^{-1}$. Mohr–Coulomb plasticity limits stresses in the crust and whereas Peierls plasticity limits differential stresses in the mantle lithosphere.](image-url)

| Table 1 | Applied rock-physical and rheological parameters. |
| --- | --- | --- | --- |
| | Upper crust (wet quartzite) | Lower crust, weak (diabase) | Mantle, wet (wet olivine) | Mantle, dry (dry olivine) |
| Elastic shear module (Pa) | $1 \times 10^{10}$ | $1 \times 10^{10}$ | $1 \times 10^{10}$ | $1 \times 10^{10}$ |
| Prefactor (s$^{-1}$ Pa$^{-1}$) | $5.07 \times 10^{-18}$ | $3.2 \times 10^{-20}$ | $4.89 \times 10^{-15}$ | $4.85 \times 10^{-17}$ |
| Activation energy (J mol$^{-1}$) | $154 \times 10^5$ | $276 \times 10^5$ | $515 \times 10^5$ | $535 \times 10^5$ |
| Power law exponent | 2.3 | 3.0 | 3.5 | 3.5 |
| Density (kg m$^{-3}$) | 2800 | 2900 | 3300 | 3300 |
| Cohesion (Pa) | $1 \times 10^7$ | $1 \times 10^7$ | $1 \times 10^7$ | $1 \times 10^7$ |
| Friction angle (deg) | 30 | 30 | 30 | 30 |
| Thermal conductivity (W m$^{-1}$ K$^{-1}$) | 2.5 | 2.5 | 3.0 | 3.0 |
| Heat capacity (J kg$^{-1}$ K$^{-1}$) | 1050 | 1050 | 1050 | 1050 |
| Radioactive heat (W m$^{-3}$) | $1 \times 10^{-6}$ | $5 \times 10^{-7}$ | 0 | 0 |
| Thermal expansivity (K$^{-1}$) | $3.2 \times 10^{-5}$ | $3.2 \times 10^{-5}$ | $3.2 \times 10^{-5}$ | $3.2 \times 10^{-5}$ |

2.3. 1-D semi-analytical model

In order to obtain a better insight in the results of the 2-D simulations, we have also developed a 1-D model, which has exactly the same stratification as the 2-D setup but assumes that the lithosphere is laterally homogeneous and deformed by pure shear extension only. The composition and boundary conditions are laterally homogeneous. The force equilibrium gives:

\[ \frac{\partial \sigma_{zz}}{\partial z} + \rho g = 0 \]  \hspace{1cm} (6)

The model is first considered viscoelastic:

\[ \dot{\varepsilon}_{xx} = \frac{\tau_{xx}}{2\mu} + \frac{1}{2G} \frac{\partial \tau_{xx}}{\partial t} \]

\[ \tau_{zz} = -\tau_{xx}, \quad \dot{\varepsilon}_{zz} = -\dot{\varepsilon}_{xx} \]

By discretizing the time term of the equation implicitly, the stress for the new time step at given strain rate is

\[ \tau_{xx}^{new} = \frac{2\mu G \Delta t \dot{\varepsilon}_{xx} + \mu \tau_{xx}^{old}}{\mu + G \Delta t} \]

If the stresses are larger than the yielding stresses, plasticity is applied. The plasticity applied in the crust is Mohr–Coulomb yield function (see Eq. (4)). In the 1-D model, when yielding occurs, the stresses are set to be the yield stresses, i.e.

\[ \sigma_{yz} = -\sigma_{zz} \]

\[ R_y^M = \frac{\sigma_{yy} \sin(\phi) + C \cos(\phi)}{1 + \sin(\phi)} \]

\[ \sigma_{yy}^y = \sigma_{zz}^y - 2R_y^M \]

\[ \sigma_{xx} = -\sigma_{yy}^y \]  \hspace{1cm} (7)

where \( R_y^M \) is the critical radius of Mohr circle of yielding and \( \sigma_{yy}^y \) and \( \sigma_{zz}^y \) are the critical first and third principal stresses of yielding, respectively. \( \sigma_{zz} \) is zero and \( \sigma_{xx} \) is given by Eq. (6). In the mantle, if the differential stresses are larger than 200 MPa, Peierls plasticity (Eq. (5)) is applied.

3. 2-D results

Over 300 2-D simulations are performed by systematically varying lithosphere thermal age and extensional strain rate, which results in three end members of deformation modes: no localization, symmetric localization and asymmetric localization (see Figs. 2–4).

An example of a simulation in which localization does not occur ("no localization mode") is shown in Fig. 2 (simu592, 10 \( ^{14} \) s\(^{-1}, 200 \) Ma). It has a thermal age of 200 Ma and an extension strain rate of 10 \( ^{14} \) s\(^{-1}. The whole model results in a pure-shear thinning until more than 100% extension. The deformation is homogeneous during extension. The temperature at the Moho, which indicates the potential for thermal localization, is nearly constant which indicates that no localized shear zone is initiated.

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![Fig. 2](https://via.placeholder.com/150)

**Fig. 2.** Snapshots of a simulation in the no localization mode. The model with a (young) thermal age of 200 Ma extends at a strain rate of 10 \( ^{14} \) s\(^{-1}. Strain rate fields at different deformation stages are shown in the first three plots. Moho temperature and a criteria (blue line) that indicates the onset of localization if the temperature is locally 50 degrees larger than the average are also shown above each plot of strain rate field. The bottom plot shows the composition field and isotherms.
A model with the same parameters but with an older thermal age (400 Ma) results in symmetric localization (simu600, $10^{-14} \text{s}^{-1}$, 400 Ma) (Fig. 3). The deformation in this model is relatively homogeneous until $\sim 25\%$ strain. After 50$\%$ strain, the lithosphere undergoes necking at the location where a set of symmetric shear zone develop. The temperature at the Moho depth arises locally about 50 deg up in the shear zone.

If the thermal age of the lithosphere is further increased (1000 Ma) with the same overall parameters, asymmetric localization occurs (simu591, $10^{-14} \text{s}^{-1}$, 1000 Ma) (Fig. 4). Due to the randomness of tracers in our models, the localized shear zone is randomly located. As seen in Fig. 4, the shear zone is observable at $\sim 25\%$ strain, and asymmetry is significantly developed at $\sim 30\%$ strain, which is notable in both strain rate and composition fields. Contour plots of temperature indicate that the temperature in the shear zone, which is fully asymmetric, is more than 100 deg higher than the surrounding rocks. The maximum amplitude of localization appears at the upper most mantle.

In our study, localization arises from thermal weakening associated with shear heating. As a result, the temperature in the localized shear zone is larger than that of the surrounding rocks. We can thus distinguish localization from no localization models in an automated manner by using the Moho temperature as a criteria. Here, we consider models to be localized if the Moho temperature is locally >50 deg higher than the average (see Figs. 2–4). To distinguish symmetric versus asymmetric localization, however, we use an empirical method based on the 2-D composition and strain rate distribution. Only cases with well-defined narrow shear zones (Fig. 4) across the lithosphere are treated as asymmetric localization (Fig. 3).

4. Semi-analytical model

In this section we develop a semi-analytical model based on the scaling law derived by Kaus and Podladchikov (2006). Following the work of Crameri and Kaus (2010), we define localization number as

$$I_{loc} = \frac{\bar{\Delta}L}{L_{com}} \sqrt{\frac{\bar{H}_{eff}}{K\rho c_p}} \frac{\Delta L}{L_{com}} \sqrt{\frac{\bar{\Delta}T_{max}}{2K\rho c_p}},$$

which predicts localization when $I_{loc} > 1$. All parameters are known in the 1-D code described in Section 2.2 except the length scale $\Delta L$. The onset of localization is independent on the shear modulus and the initial stress, but is a strong function of the length scale $\Delta L$ (Kaus and Podladchikov, 2006). In the work of Kaus and Podladchikov (2006), $\Delta L$ is simply the radius of initial heterogeneity with a temperature step. However, as we do not introduce any initial thermal or rheological heterogeneity of significant size, the physical meaning of the length scale $\Delta L$ in our work is unclear. In the study of Crameri and Kaus (2010) the thickness of plastic layer is used as the characteristic length scale $\Delta L$. Here, we take a different approach. Since the mechanism of localization we consider in this work is the coupling of shear heating and temperature-dependent viscosity, here we derive the length scale $\Delta L$ based on the additional temperature increase induced by shear heating, compared to a model without shear heating.

First, we consider a simple 1-D heat diffusion model with an initial temperature step $\Delta T$ in the center with a width $\Delta L$ (see Fig. 5a). Temperature diffuses until the maximum temperature becomes slightly smaller than $\Delta T$. As shown in the figure, the initial perturbation width $\Delta L$ is approximately equal to the part where temperature is greater than $\Delta T/2$, which is called “peak width” hereafter.
Thus one can recover the initial perturbation width $\Delta L$ from a given temperature profile. Similarly, in the current study, the temperature differs from steady state as $T_{\text{diff}} = T_{\text{SH}} - T_{\text{noSH}}$, where $T_{\text{SH}}$ is the temperature profile computed in a 1-D model with shear heating, and $T_{\text{noSH}}$ that in a model without ($H_s = 0$ in Eq. (1)). Typically a single peak is located in the mantle lithosphere where the stress is

Fig. 4. Asymmetric localization mode. Conditions are the same as in the models of Figs. 2 and 3, but the thermal age of the lithosphere was increased to 1000 Ma. The Moho temperature shows the occurrence of localization, with the composition field indicating that the shear zone is completely asymmetric.

Fig. 5. Definition of length scale $R$. (Left) Relationship between initial temperature heterogeneity (dash line) and diffused temperature profile (solid line) in a diffusion model. (Right) Differential temperature between a 1-D model with and the one without shear heating, with peak width defined.
largest. Yet in some cases, for example in a warm lithosphere, there will be another smaller peak in the upper crust and this leads to an over-estimating of $\Delta L$. Therefore, we use $\max(T_{\text{diff}})/2 + 4$ as the temperature criteria to compute the length scale $\Delta L$. Tests indicate that the amount of additional 4 deg is sufficient to overcome this problem while it is negligible once the single peak is dominating.

With the characteristic length scale $\Delta L$ being defined and the stress profile (Fig. 6a) being calculated, we can compute $I_{\text{loc}}$ as a function of depth for a given setup (Fig. 6b). The maximum of $I_{\text{loc}}$ appears in the upper most mantle, consisting with 2-D simulations which show that localization is typically initiated below the Moho (e.g. Fig. 4). If the maximum of $I_{\text{loc}}$ is greater than 1, localization is expected to occur. As we use a viscoelastic rheology, $I_{\text{loc}}$ evolves with time, and a typical evolution of the maximum of $I_{\text{loc}}$ for non-localization, symmetric and asymmetric localization as a function of strain is shown in Fig. 6d. Localization is predicted to occur for the setups in Figs. 3 and 4, consistent with the numerical results.

Moreover, we can also distinguish different deformation modes of symmetric and asymmetric localization. The evolution of localization number provides insights into the mechanism of localization. In the setup (c) (asymmetric localization), $I_{\text{loc}}$ first increases rapidly and then decreases after around 15% strain (see Fig. 6d). Although the strain when the localization number gets maximum is somewhat earlier than that of 2-D simulation (see Fig. 4), they are consistent to the first order. The decreasing of localization number may be explained as following. In setup (c) with an old thermal age, in which the strength of the lithosphere is quite strong, the rate of thermal energy generated by shear heating is much faster than thermal diffusion and therefore the energy strongly accumulates in a narrow shear band. With the increasing of the maximum of $T_{\text{diff}}$, the width of heterogeneity, as defined by the peak width, becomes narrower (see Fig. 6c). Thus the localization number reduces when $T_{\text{diff}}$ is arising. In this setup, the maximum of $T_{\text{diff}}$ exceeds 100 °C (Fig. 6e), reduces the viscosity by a factor of 10 or more, which was previously demonstrated to be

![Fig. 6. Stress profile (a) and localization number (b) calculated by the semi-analytical model for the setups of Figs. 2–4 at 15% of total strain. The black line in (b) shows the critical value for the localization number. Evolution of the length scale $R$ (c), the localization number $I_{\text{loc}}$ (d) and the differential temperature (e) are shown for the setups of Figs. 2–4. The critical value (dash line in (d)) for the onset of localization is illustrated. Panels (f) and (g) illustrate the effective and power law viscosities (solid lines) for the setups of symmetric localization and asymmetric localization, respectively. As a comparison, evolutions of viscosities for the same setups but without shear heating are shown as dash lines. In the case of asymmetric localization, the power law viscosity drops by more than an order of magnitude, whereas it is much less for symmetric localization.](image-url)
sufficient to result in asymmetric localization (Huismans and Beaumont, 2003). In contrast, in the symmetric localization setup, the localization number increases continuously until the final state. The rate of heat production in this simulation is only slightly larger than the rate of heat conduction. Therefore the width of the peak grows continuously due to the diffusion of $T_{\text{off}}$ at the same time when the energy of shear heating accumulates in the shear zone. $I_{\text{loc}}$ increases continuously and gets significantly large when the strain is more than 50%. This is well consistent with 2-D simulations (see Fig. 3) that necking of the lithosphere occurs after 50% of strain. The maximum of $T_{\text{off}}$ in this case is around 50 °C, that reduces the viscosity by a factor of around 3. Combining with the fact that the localization number is greater than 1, this predicts a symmetric localized zone, in accordance with 2-D numerical results.

A further probe of viscosity evolution that compares the 1-D model with shear heating (solid lines) with the one without (dash lines) is performed for the setups in which localization develops (see Fig. 6f and g for setups (b) and (c), respectively). We show the temporal evolution of both the effective and power law viscosity in the mantle lithosphere, where the effective viscosity includes the effect of low-temperature plasticity. The viscosities in the setups without shear heating remain nearly constant, which indicates that the self-softening effect without shear heating is insufficient to generate localization. In contrast, the viscosities decrease significantly once shear heating is introduced. Whereas the power law viscosity in the setup of symmetric localization reduces by a factor of around 3, a factor of $>10$ for the reduction of power law viscosity occurs in the setup of asymmetric localization. Yet the reduction of effective viscosity, in fact the Peierls viscosity here whose rheology does not strongly depend on temperature, is only minor, which is caused by the fact that stresses do not change dramatically in the 1-D model, which is why the mantle lithosphere remains in the Peierls plasticity deformation regime. In 2-D models, however, stress drops very rapidly once the shear zone is initiated (Schmalholz et al., 2009). As a result, the shear-zone deforms in the power law creep regime rather then in the Peierls regime. The power law viscosity reduction of a factor 10 or more is in this case sufficient to induce asymmetric localization (Huismans et al., 2005).

Based on the discussion above, we can thus use the 1-D models to compute a ‘phase-diagram’ of localization for the standard model (Fig. 7a) and the model with dry olivine mantle (Fig. 7b). Since the thermal age affects the strength of the lithosphere via temperature, which is in reality a key factor of viscosity, we convert thermal ages to temperatures at the depth of 200 km on the right side in Fig. 7. The boundary between no localization and localization is given by $I_{\text{loc}} = 1$. The boundary between symmetric and asymmetric localization is computed by recording the maximum drop of power law viscosity due to shear heating. Asymmetric localization occurs in models in which power law viscosity drops by a factor of 10 or more in the 1-D models. In addition, we have performed 2-D numerical simulations to test the 1-D predictions. The 2-D simulations we have performed cover Earth-like strain rate ranging from $10^{-16}$ to $10^{-13}$ s$^{-1}$. Yet convection is initiated in the asthenosphere in 2-D models with a background strain rate smaller than $5 \times 10^{-15}$ s$^{-1}$. In this case, most of the computation is spent on resolving the convection and the simulations become very time consuming. Therefore most of our 2-D simulation results cover relatively fast strain rates. Comparisons of such phase-diagrams and 2-D results are plotted in Fig. 7 for both the standard setup and the one with dry olivine mantle. The different mechanical modes of deformation predicted by the semi-analytical model are highlighted as shaded areas. Insets of 2-D simulations for the setups discussed above illustrate different modes of deformation (Fig. 7). There is a good agreement between 1-D and 2-D results, which thus demonstrates that the semi-analytical model is capable of predicting the onset of symmetric/asymmetric localization.

5. Applications

5.1. Stability of cratons

A number of studies addressed the longevity of cratons to the chemical buoyancy due to high degrees of melt-depletion and the high viscosity imparted by the low temperatures (Jordan, 1975, 1978; Pollack, 1986; Doin et al., 1997; Shapiro et al., 1999; Sengor, 1999). However, geodynamic calculations suggested that they are relatively insufficient to prevent the cratonic root from being recycled during convection over billions of years (Lenardic et al., 2000, 2003; Sleep, 2003; O’Neill et al., 2008). A recent analysis showed that cratons are more likely to be generated if the Rayleigh number of the mantle is larger, such as was the case in the Archean (Cooper and Conrad, 2009). The decrease of mantle Rayleigh-number with time increases basal tractions, which makes it more likely for the craton to be destroyed. Yet, as soon as a craton is surrounded by weak zones, the stresses associated with mantle convection are buffered and the craton is prevented from being further deformed (Lenardic et al., 2000, 2003). Such weak zones could result from either symmetric or asymmetric localization of the lithosphere under extension, and additionally contribute to the longevity of cratons.

Our semi-analytical model predicts a wide range of parameter space for localization for both the standard model and the model with dry olivine mantle. On one hand, at a certain extension strain rate, a cold thermal state (old thermal age) tends to develop localization, while the lithosphere under a hot thermal state (young thermal age) is more likely to deform without localization. The conditions for localization predicted in our study are in good agreements with the initial thermal condition that forms narrow rifts (Buck, 1991). Buck (1991) pointed out that none of narrow rifts were formed in a hot lithosphere (with a heat flow higher than 60–70 mW m$^{-2}$). Therefore a sufficient cold thermal state is required to keep cratons stable. On the other hand, with a certain thermal age, faster deformation is favored to initiate localization. For cratonic thermal ages (>400 Ma), a major part of Earth-like extension strain rate locates in the region of localization, with dry lithosphere developing localization easier than wet lithosphere. It implies that most cratons can form narrow rifts and keep stable. In other words, it is more likely that cratons may deform uniformly without localization under very slow extension or in a water-rich condition.

5.2. Water in the mantle lithosphere

The strength of olivine is strongly dependent on the content of water (Chopra and Paterson, 1984; Karato et al., 1986; Mackwell et al., 1985; Hirth and Kohlstedt, 1996; Kohlstedt, 2006). The water content in the upper mantle may vary regionally (Huang et al., 2005; Katayama et al., 2005; Ichiki et al., 2006; Peslier et al., 2010). A recent study by Karato (2011) has suggested high water content (up to 1 wt%) in the mantle transition zone, with eastern China significantly higher than other continental regions. Some authors (e.g. Zhu and Zheng, 2009) suggested that the destruction of NCC may be related to water content in the mantle. The NCC is one of the most striking examples for the destruction of cratons, which was proposed to have formed in the Paleoproterozoic age by the amalgamation of two Archean blocks, the western (or “Ordo”) and eastern blocks, along the Trans-North China orogen (or “Central Block”) (e.g. Zhao et al., 2001). Later in the late Mesozoic and Cenozoic, the eastern part underwent from having a thick (~200 km) lithosphere to its current much thinner lithosphere (60–80 km) (Griffin et al., 1998; Chen et al., 2006a, 2008; Zheng et al., 2008, 2009; Zhao et al., 2009). The western part however
remains stable, with a lithosphere thicker than 200 km (Chen et al., 2009; Zhao et al., 2009; Chen, 2010). A narrow rift (Shaanxi-Shanxi Rift on the order of 100 km wide) in the Central Block separates the NCC into the stable western block and the destructed eastern block (Zhao and Xue, 2010). The eastern NCC is believed to have been affected by an extensional stress field, as evidenced by a series of half-graben basins as well as seismic anisotropy patterns (Ren et al., 2002; Zhao and Zheng, 2005; Zhao et al., 2007, 2008). It has formed a basin (Bohai Bay Basin) with a maximum width of about 450 km (Chen et al., 2006b; Zhao and Xue, 2010). Receiver function studies (Chen et al., 2006b; Zheng et al., 2007, 2008, 2009) showed a thinned (~30 km) crust with small thickness variations in the eastern NCC compared to a thicker (~40 km) crust in the western NCC. It thus suggests that the eastern NCC has deformed in a “wide rifting” mode. A seismic tomography study beneath the northeast Asia shows a stagnant slab in the mantle transition zone, whose front end is located below the boundary of the eastern and western NCC (Zhao and Ohtani, 2009; Huang and Zhao, 2006). It is thus possible that the lithosphere of the eastern NCC is water-rich due to the dehydration of the subducted oceanic slab compared to the western NCC.

Here, we apply our semi-analytical model to both the eastern and western NCC. Since the western NCC is stable, it is reasonable to take its present-day stratification as the pre-extension model for the eastern NCC. Our eastern/western NCC model is defined having a 200 km lithosphere, including a 25 km upper crust and a 15 km lower crust. The rheologies for the upper crust and lower crust are the same as that in the standard model. The rheologies for the mantle, however, are different. “Dry olivine” is applied in the western NCC model while “wet olivine” is used in the eastern NCC model. The results for the occurrence of localization show that the thermal age required for localization in the western NCC model at a certain background strain rate (say ~10^{-15} s^{-1}) is generally younger than that in the eastern NCC model (Fig. 8a). This implies the western NCC is easier to develop localization and form narrow rifts, which keeps the western NCC stable. In contrast, the eastern

![Fig. 7. Prediction results of the semi-analytical model for the standard model (a) and the model with dry olivine mantle (b). Data points indicate 2-D simulation results. Insets show three deformation modes for the setups in Figs. 2–4. Temperatures at the depth of 200 km for thermal ages on the left side are shown on the right side, respectively.](image-url)
NCC tends to deform without localization. The shaded area in Fig. 8a indicates the conditions under which the western NCC is stable and the eastern NCC is destructed. For the NCC prior to extension, in which the thermal age is assumed to be 500 Ma, the temperature is about 1300 °C at the base of the 200-km thick lithosphere. According to the prediction, our model constraints the maximum extension strain rate of NCC to be $1 \times 10^{-15} \text{s}^{-1}$. Considering the thinning of crustal thickness from 40 to 30 km in the eastern NCC, such a strain rate predicts a minimum extension period of $24 \times 10^{6} \text{Ma}$. This coincides in the first order with the duration of large-scale gold mineralization (Yang et al., 2003) and magmatic activity (Wu et al., 2005) in this region.

As the next step, we combine our semi-analytical model with a rheology of wet olivine which is a function of water concentration that is introduced by Li et al. (2008). We take all parameters from Li's study except ignoring activation volume due to the lack of pressure dependency in our model. The flow law of olivine is given as (Mei and Kohlstedt, 2000)

$$\dot{\varepsilon} = A_{\text{cre}} \tau^{n_{1}} f_{\text{H}_{2}\text{O}} \exp \left(-\frac{Q_{\text{cre}}}{RT}\right)$$

with $A_{\text{cre}} = 1600 \text{MPa}^{-(n_{1}+r)} \text{s}^{-1}$, $n_{1} = 3.5$, $r = 1.2$, $Q_{\text{cre}} = 530 \text{kJ mol}^{-1}$; $\dot{\varepsilon}$ and $\tau$ are strain rate and shear stress, respectively.

$f_{\text{H}_{2}\text{O}} = \exp(c_{0} + c_{1} \ln C_{\text{OH}} + c_{2} \ln^{2} C_{\text{OH}} + c_{3} \ln^{3} C_{\text{OH}})$

is water fugacity, with $c_{0} = -7.9859$, $c_{1} = 4.3559$, $c_{2} = -0.5742$, $c_{3} = 0.0337$, and $f_{\text{H}_{2}\text{O}}$ in MPa and $C_{\text{OH}}$ is water concentration in H/10^{6}Si.

Fig. 8. (a) Onset of lithospheric-scale localization applied to the NCC. The solid line indicates the boundary of localization for the western NCC, and the dash line for the eastern NCC. The shading area figures out the parameter space when the western NCC develops localization whereas the eastern NCC does not. The mantle in the western NCC is assumed to have a dry olivine mantle rheology, whereas the eastern NCC is wet. (b) Onset of localization versus extension strain rate and water concentration in olivine for the setup of NCC which is assumed to have an initial thermal age of 500 Ma.
By fixing thermal age at 500 Ma for cratonic condition, and varying extension strain rate and water concentration, our prediction suggests that the critical minimum water concentration required to let the Eastern NCC deform without localization depends on the extension strain rate (Fig. 8b). For small strain rates, less water is required for the Eastern NCC to keep remaining in the “no localization” phase. Once the extension strain rate is determined, the critical amount of water can then be obtained. Given the maximum extension strain rate for the Eastern NCC, for example, at $10^{-15} \text{s}^{-1}$, at most $\sim 3000 \text{H/100Si}$ of water is sufficient to reduce the strength of the lithosphere and make it deform in the manner of wide-rift. This is comparable with water content in the mantle transition zone or subduction zones suggested by other studies (e.g. Dixon et al., 2004; Karato, 2011).

However, whereas the crust in the eastern NCC thins from $\sim 40$ to $\sim 30$ km, the lithosphere has much larger thinning from $\sim 200$ to $\sim 100$ km, due to differential thinning (e.g. McKenzie, 1978; Steckler, 1985; White and McKenzie, 1988), which is not well predicted in our model. Some other mechanisms might thus also partly contribute to the thinning of the lithosphere. Hydration alone, for example, could weaken the lithosphere and initiate the thinning of the mantle lithosphere (Li et al., 2008). It has been suggested (Komiya and Maruyama, 2007; Kusky et al., 2007) that the dual subduction of the Pacific and Indo-Australian plates may double the amount of water transported into the mantle transition zone under the marginal basins of the Western Pacific and under the North China Craton. Therefore, the mantle transition zone might be a reservoir of water during plate tectonics and raises a possibility to hydrate the lithosphere of the eastern NCC before extension.

5.3. Peierls strength

Only a few experiments on low temperature plasticity (Peierls plasticity) have been carried out over the past decades (Goetze, 1978; Evans and Goetze, 1979; Raterron et al., 2004; Katayama and Karato, 2008; Mei et al., 2010), and as a consequence, the flow laws are not well constrained. In order to better understand the effect of uncertainties in experimental parameters on the initiation of shear localization we compare results of a newly derived Peiers plasticity flow law (Mei et al., 2010) with the one used in this study (Goetze and Evans, 1979) (Fig. 9), which can be evaluated in a rapid manner using our 1-D models. The flow law of the newly derived Peierls plasticity by Mei et al. (2010) is described as

$$\dot{\varepsilon} = A_p \sigma^\alpha \exp \left( -\frac{E_p(0)}{RT} \right) \left( 1 - \frac{\sigma}{\sigma_f} \right)^r$$

with $A_p = 1.4 \times 10^{-7} \text{s}^{-1} \text{MPa}^{-2}$, $E_p(0) = 320 \text{km mol}^{-1}$ and $\sigma_f = 5.9 \text{GPa}$. A plot of “Christmas Tree” indicates that the Peierls plasticity by Goetze and Evans (1979) has higher strength in the mantle (Fig. 9a). The prediction results (Fig. 9b) illustrate that the general trends remain the same but that Peierls plasticity affects both symmetric and asymmetric localization in parameter space with old thermal ages (>400 Ma), in which the temperature of the lithosphere is low enough to meet the condition for “low-temperature” plasticity. Both symmetric and asymmetric localization is easier to initiate for the model with Peierls plasticity by Goetze and Evans (1979). In particular, the slower of the extension strain rate, the more important of the influence. Therefore simulations with a higher Peierls plasticity are more likely to develop localized zones, and suggest increasing possibility of stabilizing cratons.

6. Conclusions

We perform a systematic study on the conditions of strain localization in a thick lithosphere under extension, in which localization is caused by shear-heating related thermal weakening. In systematic 2-D models, we observe three end members of deformation of the lithosphere under extension, i.e. no localization, symmetric localization and asymmetric localization. A semi-analytical method is developed that gives insights in the underlying physics and helps to constrain the conditions for each of those end members. Sharp boundaries exist between different modes of deformation. We conclude that

(1) A cold thermal state tends to develop lithospheric localization in comparison with a hot thermal state.
(2) Faster deformation and colder thermal state are required to form lithospheric asymmetric localization in comparison with symmetric localization in our standard setup.
(3) Mantle lithosphere with dry olivine rheology promotes the occurrence of localization.

(4) Model Peierls’s plasticity that results in larger differential stresses (Goetz and Evans, 1979) predicts localization to occur easier.

Applications to the destruction of cratons and continental breakup are discussed. As localization forms narrow rifts and projects cartons from being destructed, higher strength of the lithosphere, e.g. stronger rheology and colder thermal state, as well as faster deformation increases the likelihood that a localized lithospheric rift occurs, which shields the remaining pieces of the craton from subsequent deformation. Lithospheres with a weaker rheology instead deform in a homogeneous manner similar to the formation of wide rifts.

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